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POLYMETAMORPHISM OF THE CRYSTALLINE BASEMENT OF THE SOMOGY—DRAVA BASIN (SOUTHWESTERN TRANSDANUBIA, HUNGARY)

P. ÁRKAI

ABSTRACT

The crystalline basement of the Neogene depression of the Somogy—Drava Basin (SW-Transdanubia, Hungary) consists of medium-grade (almandine-amphibolite facies) polymetamorphic formations overprinted by a very low- and low-grade (anchi-, epizonal) retrograde, partly cataclastic metamorphism: gneiss, mica schist, amphibolite, as well as mylonite and blastomylonite.

Having applied the petrological and geochemical methods of lithofacies reconstruction the gneiss — mica schist and the mylonite — blastomylonite groups developed from them proved to be of sedimentary (para) origin: they were formed from carbonate-free or carbonate-poor pelitic-psammitic sediments. The amphibolite originated from basic igneous rock.

On the basis of the mineral-paragenetic, petrotextural and structural as well as geothermometric and geobarometric characteristics the following relative chronological succession of metamorphic events was determined:

A) The oldest one is a medium-grade (almandine-amphibolite facies), medium pressure (Barrovian) regional metamorphism with a geothermal gradient of 17 to 27 °C/km. Its temperature and pressure were 510 to 600 °C and 5.9 to 8.9 kbar, respectively. It was locally followed by

B) an andalusite-type (low pressure range) medium-grade (amphibolite facies) metamorphism with a gradient >34 °C/km and by

C) a low temperature (<450 °C), predominantly low pressure anchi-, epizonal retrograde, locally cataclastic metamorphism.

As to our recent knowledge — based on Alpine and Carpathian analogies — different hypothetic geochronological modells can be established, e. g.

— *A)* Caledonian, *B)* Hercynian, *C)* Hercynian and/or Alpine:

— *A)* older Hercynian, *B)* younger Hercynian and *C)* younger Hercynian and/or Alpine:

— *A)* Dalslandian (Early Baikalian), *B)* Hercynian, *C)* Hercynian and/or Alpine.

The mineral assemblages formed by weathering and by low temperature retrograde metamorphism were distinguished by means of the clay mineral associations and by the illite crystallinity.

INTRODUCTION

Considering its geographic position, the Somogy—Dráva Basin links the Pannonian Basin with the Southern Alps and Internal Dinarides. This depression filled with a maximally 5.000 m thick sedimentary pile belongs to the Transdanubian Neogene Basin System. The aims of the present petrological and geochemical examinations of its crystalline basement are: the determination of the origin of these much debated polymetamorphic formations, the distinction and characterization of their metamorphic events and weathering processes. The investigations intend to contribute to the hydrocarbon prognostics of the basin on the one hand, and to the correlation of the metamorphic events of the afore-mentioned great geographic-geological units, on the other. The metamorphic-petrogenetic research of the core samples deriving from hydrocarbon exploratory wells was initiated and supported by the National Oil and Gas Industrial Trust and by its affiliated firm, the Oil and Gas Mining Enterprise.

GEOLOGY, PREVIOUS DATA

The geological and tectonic situation of the basement of the Somogy—Drava Basin is shown in *Fig. 1a.* The northern part of the basin belongs to the so called Igal—Bükk Alpine mobile belt extending between the Balaton and Zagreb—Hernád lineaments. As to the former ideas this belt might from the southwestern paleogeo-

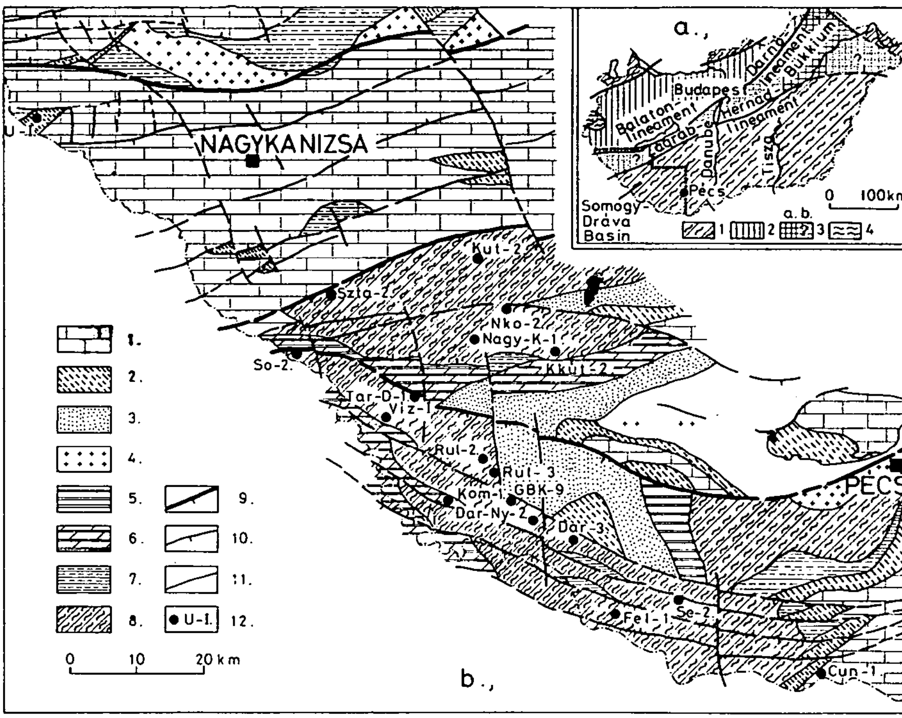


Fig. 1. a: Geological-tectonic position of the Somogy—Drava Basin.
Legend: 1) medium-grade polymetamorphic crystalline basement (pre-Hercynian?-Hercynian-Alpine?) locally with synkinematic Hercynian granitoid formations 2) Hercynian very low and low-grade metamorphic Early Paleozoic with post-kinematic Hercynian granitoid intrusions with non-metamorphic Late Paleozoic and Mesozoic (Transdanubian Central Mountains and the Little Plain) 3) Alpine very low- and low-grade metamorphic Paleozoic and Mesozoic (Bükkium, Igal—Bükk mobile belt): *a)* evidenced by investigations *b)* presumed *4)* Alpine low-grade metamorphic Mesozoic (Kőszeg Mountains, Penninic unit).

b: Geological and tectonic sketch of the pre-Tertiary basement of the Somogy—Drava Basin after BARDÓCZ (1973—1982). Simplified with sampling localities of metamorphic rocks. Legend: *1)* Mesozoic *2)* Permian partly Permo-Triassic *3)* Carboniferous *4)* granitoid rocks *5)* Early Paleozoic in general *6)* Devonian (?) *7)* Silurian (?) *8)* polymetamorphic crystalline basement (pre-Hercynian?-Hercynian-Alpine?) *9)* main compressive tectonic surfaces *10)* subordinate compression surfaces *11)* fault *12)* studied boreholes.

graphic ocnnection of the Bükkium characterized by South-Alpine, Dinaric affinity ("eugeosyncline", WEIN, 1969). However, the existence of this belt as a paleogeographic unit has not been proved [BALOGH, 1981, personal communication, KOVÁCS, 1982], and recently the present position of the Bükkium is explained by horizontal micro-plate movements along the Zagreb—Hernád lineament [KOVÁCS, 1982]. The stratigraphic, paleontological and lithofacies correlations of the Paleozoic and Mesozoic formations of the Bükkium and that of the southwestern part of the Igal—Bükk Belt are lacking. Traces of the Alpine (Cretaceous) regional metamorphism demonstrated in the Paleozoic and Mesozoic formations of the Bükkium [ÁRKAI, 1973, 1983, ÁRKAI, HORVÁTH and TÓTH, 1981] are unknown in the southwestern part of the belt belonging to the Somogy—Dráva Basin.

South of the Zagreb—Hernád lineament the crystalline basement is built up by pre-Hercynian (?) — Hercynian — Alpine (?) polymetamorphic formations of unclear evolution, locally accompanied by most likely Hercynian granitization (Mecsek Mountains, Danube—Tisza Interfluve).

The simplified variant of the map of the pre-Tertiary basement of the Somogy—Drava Basin constructed by BARDÓCZ (1973—1982) is shown in *Fig. 1b*. The basement units are bordered by compressive faults (overthrusts?) with strikes changing from NE—SW to NW—SE. The younger faults mostly perpendicular to the compressive ones are disjunctive.

According to the first petrographic description of the metamorphic basement of the Drava Basin [SZEPESHÁZY, 1958, 1959] most of the rocks having granite-like composition are "unequilibrated" metamorphites: tectonites, with different mylonite and phyllonite types depending on the intensity of the cataclastic metamorphism.

In the first petrogenetic synthesis of the SW-Transdanubian crystalline basement [BALÁZS, 1968] these polymetamorphites were believed to be ortho-rocks. The primary granitoid rocks were metamorphized in the Precambrian (Late Baikalian?) tectonocycle (meso- or occasionally kata-zonal gneiss formation). The epizonal mylonitization is presumably of Variscan age.

Essentially this synthesis can be found in the explanatory text of the map of metamorphites of the Carpatho—Balkan—Dinaride area [SZÁDECZKY-KARDOSS, JUHÁSZ, BALÁZS *et al.*, 1969] as well as in the map itself [SZÁDECZKY-KARDOSS, ÁRKAI *et al.*, 1976], with some modifications: the premetamorphic rocks are of Proterozoic age, an uncertain Early Baikalian metamorphism was followed by a Variscan epidote-amphibolite facies metamorphism and by an Alpine retrograde greenschist facies metamorphism. The anatexis (synkinematic granitization) observed in the polymetamorphic formations of the Mecsek Mountains and of the Danube—Tisza Interfluve seems to be also of Variscan age.

NAGY and SZEPESHÁZY [1971] pointed out the spatial variation of the grade and intensity of the Hercynian metamorphism in the basement rocks. In certain part of the Precambrian meso-zonal gneiss, mica schist, amphibolite sequence the weaker retrogressive effect resulted in mylonitization and phyllonite formation. Elsewhere these rocks were more strongly recrystallized (blastomylonites), and locally, in the high temperature zones anatexis, granitization took place without any signs of katazonal assemblages. The common greenschist facies retrograde event often called diaphtoresis can be bound partly to the Alpine cycle, as well.

Based on feldspar studies as well as on Rb—Sr isotope geochronological data BUDA [1972] established a model for the granitoid rocks of the Mecsek Mountains and of the Danube—Tisza Interfluve presuming Precambrian (?) sedimentation, Caledonian synkinematic anatexis and Variscan late-kinematic potash metasomatism.

In the region of the Görcsöny Ridge adjoining the Drava Basin, SZEDERKÉNYI [1975] demonstrated a Barrovian metamorphic sequence with increasing metamorphic grade from SW to NE from the chlorite zone up to the sillimanite zone. He determined the main tectonic directions of the basement and distinguished a two-phase (Precambrian and Variscan) granizitation in the Mecsek Mountains.

According to the lithostratigraphic correlation of JANTSKY [1976, 1979] the almandine-amphibolite facies ultrametamorphic (granitized) sequence of the Mecsek Mountains originated from Lower Proterozoic eugeosyncline pelitic-psammitic sediments and ophiolitic magmatites by pre-Baikalian (Gothian) amphibolite facies regional metamorphism and pre-Baikalian (Dalslandian) amphibolite facies ultrametamorphism. The diaphtoresis of the sequence is Baikalian (Riphean), the younger tectonocycles were not accompanied by regional metamorphic effects. The metamorphic evolution in the basement of the Drava Basin is similar to that of the Mecsek Mountains, except the ultrametamorphism (granitization).

The critical review of LELKES–FELVÁRI, SASSI *et al.*, [1981] based mainly on the studies of SZEDERKÉNYI considers the metamorphism of the crystalline basement of the Drava Basin and Görcsöny Ridge to be pre-Upper Carboniferous. As to the generally accepted view it is partly pre-Hercynian, though the sporadic age data refer to Hercynian thermal event only. The Barrovian, medium temperature gradient metamorphism (kyanite+staurolite) was overprinted by an andalusite-cordierite type, high thermal gradient recrystallization. This latter was locally accompanied by anatectic migmatizitation. At present the ages of the two metamorphic episodes are not known exactly.

It is obvious from this short review that our knowledge on the lithofacies of the premetamorphic rocks and on the polymetamorphic events and their chronology is full of contradictions and has not cleared up so far. The new petrological and geochemical data of this paper might contribute to solve the questions above.

METHODS

Out of 15 boreholes *(Fig. 1b)* reaching the crystalline basement the complex investigation of 30 samples was carried out. In addition to the macroscopic and transmission microscopic studies X-ray diffractometric, electron microprobe, silicate-analytical and emission spectrographic methods were used.

The *X-ray diffractometric investigations* aimed:

— the determination of the qualitative and semiquantitative mineral composition, in the latter case using the direct method of NÁRAY-SZABÓ and PÉTER [1967] and also the data of BÁRDOSSY [1966, 1970], RISCHÁK and VICZIÁN [1974] as well as VICZIÁN and GHONEIM [1977]. The semiquantitative phase analysis was done by TÓTH, N. M. The results were corrected by the microscopic observations, by the solution residue data obtained by treatment with 3% HCl, and by the mineralogical recalculation of whole rock chemical analyses using also the chemical compositions of minerals determined by electron microprobe.

— The *measurement of illite crystallinity indices* referring to the structural ordering of the illite-muscovite group. The measurements were carried out in whole rock samples and in the fractions of less than 2 microns according to the methods of KÜBLER [1968, 1975] and WEAVER [1960]. The Kübler-index (=illite crystallinity= =IC=largeur de Scherrer=LS) denotes the width of the first (10 Å) basal reflection of the illite-muscovite at half height of the peak in 2Θ degrees, under standardized circumstances (this value is abbreviated as half-width). The Weaver-index (sharpness ratio) is the ratio of the peak-heights (intensities) measured on the basal reflection of the illite-muscovite at 10 and 10.5 Å (Weaver-index = I_{10} Å/$I_{10.5}$ Å).

— The *determination of the metamorphic pressure* indicating d_{060} or $6 \times d_{060} = b_0$ and the d_{002} lattice parameters of the illite-muscovite group according to SASSI [1972] and GUIDOTTI and SASSI [1976], with corrections using the (211) and (100) reflections of the quartz in the rocks as "internal standards".

The X-ray analyses listed above were made by a Philips PW—1730 type diffractometer, with the following recording conditions: CuK_α radiation, 45 kV accelerating voltage, 35 mA, proportional counter, graphite monochromator, divergency and detector slits of 1°, goniometer speeds of 2°/min and 1/2°/min, respectively, time constant of 2 sec, and paper speed of 2 cm/min.

In silicate analyses made by LEFLER, J. and HANGYÁS, GY. a Carl Zeiss AASIM type atomic absorption spectrophotometer was used, in addition to the traditional gravimetric, photometric and chromatometric methods.

Trace elements were determined by TOMSCHEY, O. by means of a Carl Zeiss PGS—2 type plane grating spectrograph. In the quantitave evaluation the combined analytical and adjusting method was used. Recording conditions were: BIG—100 AC generator, Al 9999 electrodes, 1 mm electrode distance, 15 μm split, 50/1 spark, blende: 2, amperage: 7 A, exposition time: 164 sec, collimator: 10.6, plate quality: 23D56, developing in AGFA—1 at 20 °C for 5 min.

The chemical composition of rock forming and accessory minerals were analysed by NAGY, G. and DOBOSI, G. by means of a JEOL Superprobe—733. In correction calculations the ZAF program of the JEOL firm was used. In the qualitative chemical analyses an EDAX-system was applied.

RESULTS

In Table 1 the location, number, depth interval and rock type of the investigated samples are listed. In Table 2 the semi-quantitative mineral compositions, in Table 3 the lattice parameters and the different indices characteristic of the structural ordering of the illite-muscovite group, in Table 4 the main element compositions and in Table 5 the trace element compositions are found. Table 6 contains the chemical compositions of the rock forming and accessory minerals analyzed by microprobe as well as the cation numbers per unit cells.

DISCUSSION AND CONCLUSIONS

Rock types, mineral composition

The rocks of the *gneiss group* are the most widespread (samples So—3.3, Viz—I.2,3, Tar—D—1.5,6, GBK—9.15, Dar—3.11, Fel—I.15, Se—2.16, Cun—1.14, Szta—2.20, Kut—2.16, Nagy—K—1.11, 12). These are composed of quartz, plagioclase (oligoclase), muscovite and biotite. Considerable amount of potash feldspar was found only in the sample Cun—1.14. The gneiss type containing only biotite out of the phyllosilicates is also rare (Dar—3.11). The samples So—3.3, Viz—I. 2, Dar—3.11, Se—2.16, Szta—2.21, Kut—2.16 and Nagy—K—1.11, 12 contain garnet, those of So—3.3, Dar—3.11, Kut—2.16 and Nagy—K—1.11 contain staurolite. The occurrences of kyanite (Nagy—K—1.11), sillimanite (So—3.3) and andalusite (Kut—2.16) are sporadic.

Mica schists and feldspar bearing mica schists being in close genetic and spatial relationship with the gneisses are less frequent (Viz—I. 5, 6, GB—29.2, Nagy—K—

List of the investigated rock samples from the crystalline basement of the Somogy—Drava Basin

Village	Borehole	Core	Depth (m)	Rock type
Somogyudvarhely	So—3.	3	3001.0—3010.0	gneiss
Vízvár	Víz—I	2	3287.0—3291.5	gneiss
Vízvár	Víz—I	3	3291.5—3295.0	gneiss
Vízvár	Víz—I	5	3465.0—3468.0	mica schist
Vízvár	Víz—I	6	3565.0—3566.5	mica schist
Tarany	Tar—D—1	5	3011.0—3012.0	gneiss
Tarany	Tar—D—1	6	3299.0—3300.0	gneiss
Rinyaújlak	Rúl—2	2	2720.0—2723.5	blastomylonite
Rinyaújlak	Rúl-3	2	2719.0—2723.0	phyllonitic blastomylonite
Görgeteg-Babócsa	GB—29	2	1908.0—1911.0	mica schist
Görgeteg-Babócsa	GB—29	4	2150.0—2151.0	mica schist
Görgeteg-Babócsa	GBK—9	15	2702.0—2705.0	gneiss
Darány	Dar—Ny—2	4	2697.0—2700.0	amphibolite
Darány	Dar—3	11	2976.0—2980.0	gneiss
Felsőszentmárton	Fel—I	15	3991.2—3993.2	gneiss
Sellye	Se—2	16	1798.0—1801.0	gneiss
Sellye	Se—2	18	1950.0—1952.0	gneiss
Cun	Cun—1	14	1969.5—1971.5	gneiss
Szenta	Szta—2	20	2682.0—2684.0	gneiss
Kutas	Kut—2	16	1947.0—1949.0	gneiss
Nagyatád	Nagy—K—1	10	3071.0—3073.5	mica schist
Nagyatád	Nagy—K—1	11	3220.0—3222.0	gneiss
Nagyatád	Nagy—K—1	12	3299.0—3300.0	gneiss
Nagykorpád	Nko—2	11	2065.0—2067.0	mica schist
Kadarkut	Kkút—2	4	1219.0—1222.0	mylonite
Kadarkút	Kkút—2	5	1222.0—1223.0	blastomylonite
Kadarkút	Kkút—2	8	1415.0—1417.0	blastomylonite
Kadarkút	Kkút—2	9	1531.0—1533.0	blastomylonite

1.10, Nko—2.11). Their rock forming minerals are similar to that of the gneiss group, the only difference being the lower (less than 20 weight percent) feldspar content. Their quartz/phyllosilicate ratios are rather changing. The accessory minerals are garnet (GB—29.2 and Viz—I.5) as well as straurolite (GB—29.2).

Amphibolite is subordinate in the basement (Dar—Ny—2.4). Its minerals are hornblende, plagioclase (oligoclase), quartz and small amounts of biotite, garnet and epidote.

The *mylonite* (GB—29.4, Se—2.18, Kkut—2.4) and *blastomylonite* varieties (Rul—2.2, Kkut—2.5, 8,9) were generated from the gneisses and mica schists by cataclastic (dynamic) metamorphism accompanied by a weak retrograde recrystallization which did not exceed the chlorite zone of the greenschist facies. These rock types are characterized by smaller feldspar content and by high sericite, chlorite and carbonate mineral contents as compared to the gneisses.

Pre-metamorphic lithofacies

The overwhelming majority of the gneisses, mica schists and mylonites, blastomylonites generated from them proved to be of *sedimentary* (para-) origin. They were formed presumably from carbonate-free or carbonate-poor pelitic-psammitic

TABLE 2

Semiquantitative mineral composition of the rock samples (weight percent)
(Analysed by P. ÁRKAI and M. N. TÓTH)

Sample	Quartz	Plagioclase	K-feldspar	Hornblende	Biotite	Muscovite (+illite-sericite)	Chlorite	Kaolinite	Mixed layer clay min.	Calcite	Dolomite-ankerite	Siderite	Pyrite	Hematite	Goethite	Garnet	Staurolite	Kyanite	Sillimanite	Andalusite	Epidote
So—3.3	34	32	—	—	6	17	3	2	—	—	—	5	—	1	—	tr	tr	—	tr	—	—
Víz—I.2	41	26	—	—	—	20	12	—	—	1	—	—	—	tr	—	tr	—	—	—	—	—
Víz—I.3	18	56	—	—	—	tr	22	—	—	4	—	tr	—	—	—	tr	—	—	—	—	—
Víz—I.5	14	11	—	—	15	45	14	—	—	—	tr	—	—	—	—	tr	—	—	—	—	—
Víz—I.6	64	18	—	—	—	8	10	tr	—	—	—	—	—	—	—	tr	—	—	—	—	—
Tar—D—1.5	33	18	—	—	tr	12	4	6	—	1	12	14	—	tr	—	—	—	—	—	—	—
Tar—D—1.6	39	20	—	—	tr	16	19	—	—	6	—	—	—	tr	—	tr	—	—	—	—	—
Rul—2.2	36	20	—	—	—	25	2	4	—	—	—	13	—	tr	—	—	—	—	—	—	—
Rul—3.2	49	12	1	—	—	21	—	6	—	—	—	11	—	tr	—	tr	—	—	—	—	—
GB—29.2	57	8	—	—	16	15	—	2	—	—	—	2	—	tr	—	tr	tr	—	—	—	—
GB—29.4	68	8	—	—	—	10	14	—	—	—	—	—	—	—	—	tr	—	—	—	—	—
GBK—9.15	46	38	—	—	—	10	3	—	—	3	—	tr	—	—	—	—	—	—	—	—	—
Dar—Ny—2.4	3	30	—	57	tr	—	10	—	—	—	—	—	—	—	—	tr	—	—	—	—	tr
Dar—3.11	46	31	—	—	15	—	6	—	—	1	1	—	—	—	—	tr	tr	—	—	—	—
Fel—I.15	39	55	—	—	tr	3	3	—	—	—	tr	—	—	tr	—	—	—	—	—	—	—
Se—2.16	45	15	—	—	21	2	2	15	—	—	—	—	—	—	—	tr	—	—	—	—	—
Se—2.18	44	2	—	—	—	22	tr	8	—	—	—	24	—	—	—	—	—	—	—	—	—
Cun—1.14	47	—	30	—	13	8	—	—	—	—	1	tr	—	—	1	—	—	—	—	—	—
Szta—2.20	14	33	—	—	—	19	29	—	—	5	—	—	—	tr	—	tr	—	—	—	—	—
Kut—2.16	21	21	—	—	9	26	—	8	—	—	13	—	2	—	—	tr	tr	—	—	tr	—
Nagy—K—1.10	33	—	1	—	—	49	—	5	—	—	—	11	—	1	—	—	—	—	—	—	—
Nagy—K—1.11	24	51	—	—	9	7	1	3	—	—	2	2	—	1	—	tr	tr	tr	—	—	—
Nagy—K—1.12	49	16	—	—	—	20	1	1	—	—	6	7	—	—	—	tr	—	—	—	—	—
Nko—2.11	38	11	—	—	8	30	—	5	—	—	—	—	—	8	—	—	—	—	—	—	—
Kkút—2.4	3	48	37	—	—	5	—	—	—	—	—	7	—	—	—	—	—	—	—	—	—
Kkút—2.5	27	40	—	—	—	12	18	—	—	—	—	3	—	—	—	—	—	—	—	—	—
Kkút—2.8	30	33	25	—	—	6	tr	2	—	—	—	4	—	tr	—	—	—	—	—	—	—
Kkút—2.9	38	35	16	—	—	6	3	—	—	—	—	2	—	—	—	—	—	—	—	—	—

tr = traces

TABLE 3

X-ray diffractometric structural parameters of the illite-sericite-muscovite group
(biotite-free samples)

Sample	Whole rock samples (desoriented preparates)						2 µm ⌀ fractions (oriented prep.)			
	IC°2Θ (Kübler-index)		Weaver-index		b_0 (Å)	d_{002} (Å)	IC° 2Θ (Kübler-index)		Weaver-index	
	2°/min	1/2°/min	2°/min	1/2°/min			2°/min	1/2°/min	2°/min	1/2°/min
Víz—I.2	0.220	0.150	6.60	7.30	9.012	9.964	n. d.	n. d.	n. d.	n. d.
Víz—I.6	0.172	0.097	8.14	9.75	9.055	9.952	n. d.	n. d.	n. d.	n. d.
Rul—2.2	0.274	0.175	3.84	5.25	8.995	9.952	n. d.	n. d.	n. d.	n. d.
Rul—3.2	0.209	0.136	5.07	5,77	9.003	9.946	0.568	0.446	1.98	2.35
GB—29.4	0.230	0.158	8.25	11.56	n. d.	9.958	0.274	0.127	4.96	8.93
Se—2.10	0.350	0.308	3.00	3.13	8.991	9.997	0.551	0.527	3.34	2.23
Szta—2.20	0.296	0.207	3.79	4.20	9.003	9.964	n. d.	n. d.	n. d.	n. d.
Nagy—K—1.10	0.203	0.126	6.52	8.20	8.996	9.958	0.481	0.420	3.25	2.86
Nagy—K—1.12	0.221	0.158	5.39	5.08	8.998	9.958	n. d.	n. d.	n. d.	n. d.
Kkút—2.5	0.209	0.124	9.50	10.88	9.906	9.952	0.208	0.140	6.25	2.69
Kkút—2.8	0.170	0.109	9.82	7.30	9.001	9.913	n. d.	n. d.	n. d.	n. d.
Kkút—2.9	0.132	n. d.	9.33	n. d.	9.008	10.048	n. d.	n. d.	n. d.	n. d.

n. d. = not determined

clastic sediments mixed in varying ratios. The observations evidenced the para-origin are as follows:

— detrital, rounded zircons in the samples Viz—I.2, 6, Ta—D—1.5, 6 and Cun—1.14;

— fine disperse organic matter (graphite) in the sample Nagy—K—1.11;

— the abundant occurrence of Al-rich silicates (garnet, staurolite, Al_2SiO_5 modifications) generated as a result of Al-excess characteristic of the pelitic rocks;

— the strong fluctuation of the quartz-feldspar-phyllosilicate ratios relating to the varying mixtures of fine and coarse detrital sediments;

— the lack or the subordinate quantity of potash feldspar denying the granitic origin;

— the lack of carbonate minerals in the first progressive mineral assemblages.

Having evaluated the rocks with considerable amounts of potash feldspar, the migmatitization of paragneiss can be presumed in the case of the sample Cun—1.14 containing rounded zircon grains. The mylonites Kkut—2.4—9 formed from granitoid rocks or orthogneiss. Assuming the granitoid origin the primary sedimentary starting material can not be excluded in this case either.

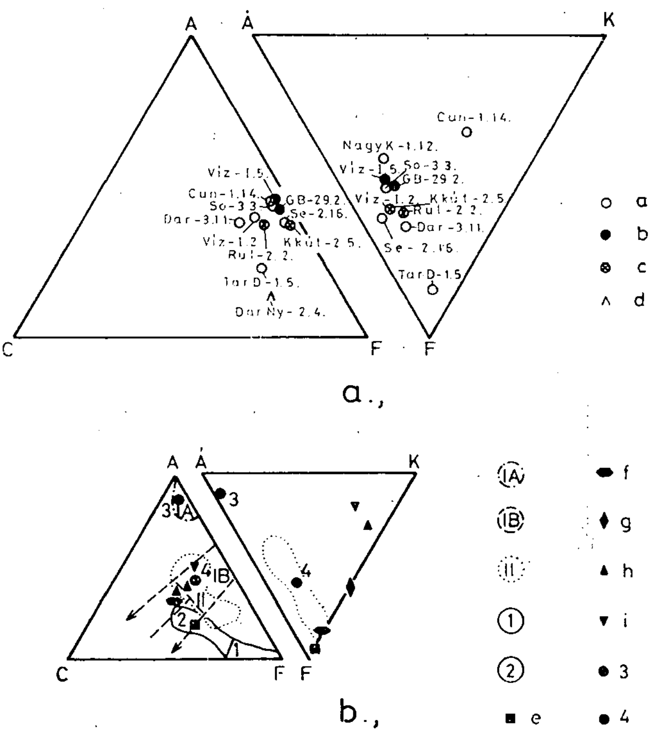


Fig. 2. a: Distribution of rocks of the crystalline basement in the ACF—ÁKF diagrams. Legend: *a)* gneiss *b)* mica schist *c)* blastomylonite *d)* amphibolite.

b: Distribution of the main rock types in the ACF—ÁKF diagrams after Winkler [1976]. Legend: IA—Al-rich clays, IB — carbonate-free or max. *35%* carbonate-bearing clays; between arrows: marls (with 35—65% carbonate content) II — greywackes, 1 — ultrabasic rocks, 2 — basic (basaltic) and andesitic rocks, e — basalt, f— tonalite, g — granodiorite, h — calc-alkaline granite, i — alkali granite, 3 — continental clay, 4 — marine clay.

Table 4

Chemical composition of the polymetamorphic samples (weight percent)
(Analysed by J. LEFLER and GY. HANGYÁS)

	So— 3. 3.	Víz— I. 2.	Víz— I. 5.	Tar D— 1. 5.	Rul— 2. 2.	GB— 29. 2.	Dar Ny—2.4.	Dar— 3. 11.	Sc— 2. 16.	Cun— 1. 11.	Cun— 1. 14.	Kut— 2. 15.	Nagy K—1.12.	K kut— 2. 5.
SiO_2	63.05	65.82	51.64	54.79	60.98	73.66	46.30	74.73	67.39	12.42	74.09	21.07	68.28	61.91
TiO_2	0.38	0.65	0.65	0.43	0.46	0.43	1.26	0.43	0.37	0.17	0.17	1.11	0.38	0.56
Al_2O_3	16.16	15.00	23.87	10.26	14.93	12.25	12.80	11.68	13.74	3.53	11.32	8.70	12.66	16.61
Fe_2O_3	2.22	3.02	2.72	4.27	2.23	0.86	4.27	1.61	1.34	0.00	1.45	3.03	3.69	2.09
FeO	3.75	2.18	5.24	5.70	2.53	3.05	9.66	2.44	4.48	0.62	1.13	7.58	1.49	3.74
MgO	2.81	3.34	4.60	5.05	3.87	2.17	10.34	1.60	3.15	6.94	1.28	14.75	1.77	3.57
MnO	0.09	0.23	0.18	0.23	0.12	0.16	0.44	0.25	0.17	0.07	0.08	0.96	0.17	0.23
CaO	0.96	1.82	1.10	3.72	1.81	0.44	6.98	1.96	0.91	29.49	0.48	15.55	1.39	0.62
Na_2O	2.21	3.01	1.89	1.33	2.16	1.10	3.86	2.75	1.64	0.14	0.64	0.64	1.60	3.59
K_2O	2.91	1.91	4.93	2.14	2.83	2.85	0.32	1.38	2.08	1.36	6.27	1.07	2.60	2.47
$+H_2O$	1.80	2.43	2.44	2.36	2.24	1.42	1.81	0.82	3.43	1.51	2.02	1.38	1.79	2.37
$-H_2O$	0.00	0.04	0.00	0.16	0.21	1.30	0.40	0.08	0.14	0.18	0.28	0.62	0.07	0.12
CO_2	2.62	0.48	1.41	9.78	4.88	0.86	0.83	1.03	0.33	6.03	0.00	21.12	4.32	1.40
P_2O_5	0.24	0.15	0.18	0.09	0.22	0.06	0.19	0.23	0.22	0.15	0.14	2.76	9.11	0.16
Total:	99.20	100.08	100.85	100.31	99.47	100.61	99.46	100.99	99.39	99.52	99.35	100.34	100.32	99.44

TABLE 5

Trace element concentrations of rock samples (ppm)
(Analysed by O. TOMSCHEY)

Sample	B	Co	Cr	Ga	Mn	Ni	Pb	Ti	V	Zn
So—3.3	15	11	150	11	367	62	5	3550	45	63
Víz—I.2	19	10	350	34	835	50	9	3500	97	58
Víz—I.3	8	3	32	9	353	15	3	2020	40	8
Víz—I.5	24	12	138	45	1060	49	9	3320	173	124
Víz—I.6	8	6	66	40	453	13	6	1660	12	11
Tar—D—1.5	20	11	112	28	2030	32	8	3230	45	63
Tar—D—1.6	16	20	135	31	673	34	12	3080	68	87
Rul—2.2	48	13	157	60	650	67	53	4030	155	88
Rul—3.2	23	8	103	34	320	27	9	2050	47	31
GB—29.2	16	8	89	27	920	21	11	3330	52	41
GB—29.4	11	—	8	11	190	—	4	630	13	12
GB—K—9.15	15	3	21	9	310	8	4	1630	24	8
Dar—Ny—2.4	17	16	95	19	2070	27	5	3620	300	34
Dar—3.11	15	—	68	17	740	15	9	2480	22	24
Fel—I.15	7	—	2	23	230	4	8	530	7	18
Se—2.16	10	—	60	8	180	9	6	1650	27	—
Se—2.18	140	7	140	30	510	94	6	3170	77	—
Cun—1.14	24	—	1	14	97	—	7	447	6	—
Szta—2.30	7	—	41	5	527	11	3	1240	17	8
Kut—2.16	70	28	182	70	1030	86	21	5830	273	85
Nagy—K—1.10	60	15	99	31	193	21	9	2320	75	35
Nagy—K—1.11	110	12	130	42	920	53	12	3030	153	83
Nagy—K—1.12	42	8	138	33	837	38	11	3920	137	42
Nko—2.11	42	8	138	31	667	33	8	3100	72	33
Kkút—2.4	15	—	4	62	893	12	14	820	60	65
Kkút—2.5	45	24	212	59	910	87	11	3370	113	101
Kkút—2.8	17	—	7	46	393	8	14	833	12	16
Kkút—2.9	11	—	5	23	237	7	9	497	10	—

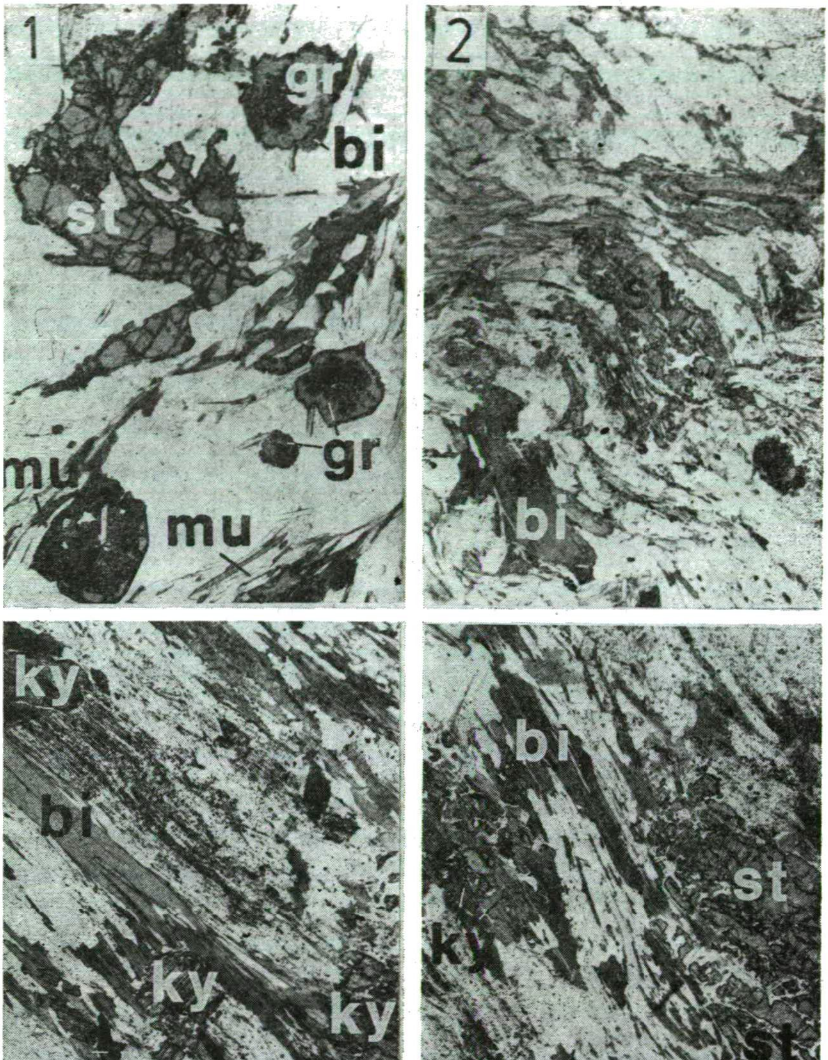
Samples GBK—9.15 and Fel—I.15 differ from the gneiss group in their high quartz+feldspar and low phyllosilicate contents. These may be the metamorphic derivates of dioritic magmatites or of feldspar-rich sandstones.

The petrochemical (main element) data support the conclusions above concerning the para origin. Plotting the petrochemical data (Table 4) in the ACF—ÁKF diagrams *(Fig. 2a)* these were compared with the average values and fields of the main sedimentary and igneous rock types determined by WINKLER [1976, 1979] shown in *Fig. 2b*. It can be stated that

— except two samples the rocks of the gneiss — mica schist and related mylonite — blastomylonite groups lie in the field of the marine carbonate-free clays and graywacke, far off the igneous trend, i.e. these are of para (pelitic-psammitic) origin. Out of the exceptions the position of the sample Tar—D—1.5 was displaced towards the C-edge due to the secondary carbonate mineral formation. In the ÁKF diagram the sample Cun—1.14 is of transitional position between the clastic sediments and granites indicating the granizitation process.

— The amphibolite sample (Dar—Ny—2.4) falls out of the marl region and corresponds to the basaltic part of the igneous differentiation trend, i. e. it is presumably of ortho origin. The projection point of the investigated amphibolite coincides also with the ortho field in the MgO—CaO—FeO discrimination diagram of WALKER *et al.* [1960].

PLATE I



The considerable fluctuations of the trace element contents of the gneiss — mica schist and mylonite — blastomylonite groups (Table 5) also indicate that these rocks were recrystallized from not homogenized detrital sediments of varying compositions within wide limits. In the trace element concentrations of the rocks mentioned above there are no remarkable, systematic differences.

The trace element ratios of the amphibolite sample fall in the ortho fields of the discrimination diagrams using the Ni—Co and Ni—Cr ratios [WALKER *et al.*, 1960] as well as the V—Cr ratio [SCHWEDER, 1968]. Taking into account the main element composition and the lack of the carbonate rocks in the basement, the amphibolite seems to be more of basic igneous than of sedimentary (marly) origin.

RECRYSTALLIZATION PHASES, METAMORPHIC EVOLUTION

The mineral assemblage of the first detectable progressive metamorphism in the gneiss — mica schist and the mylonite — blastomylonite groups consists of quartz, plagioclase (oligoclase), muscovite, biotite $\pm$ garnet, staurolite, kyanite, sillimanite and potash feldspar.

Applying the Winkler's classification [1976, 1979] this mineral assemblage involves the whole range of medium-grade metamorphism, and is assigned in the facies classification of WINKLER [1967] to the medium-pressure (Barrovian) almandine-amphibolite facies (staurolite-almandine and kyanite-almandine-muscovite subfacies). Out of the metamorphic grade indicating minerals the garnet (Plate I.1) is the most abundant. Staurolite is less frequent (Plate I.1, 2, 4). while kyanite (Plate I.3, 4) and sillimanite occur only in one-one samples. Their associations are: garnet; garnet-staurolite; garnet-staurolite-kyanite and garnet-staurolite-sillimanite.

Based on the reaction isograde system of metapelites [WINKLER, 1976, 1979] the probable temperature interval of the first progressive metamorphism is about 560—620 °C, the pressure ($P_{H_2O}=P_S$) is more than 6 kbar. (Temperature and pressure were estimated on the basis of the field bordered by the "staurolite-in" isograde and the initial anatexis, and of the field above the Al_2SiO_5 tripple point, respectively.)

The progressive metamorphic mineral assemblage of amphibolite is hornblende, plagioclase (oligoclase), quartz, biotite, garnet, epidote. The temperature of the metamorphism might have been between the Winkler's "hornblende-in" and "$An_{17}+$ $+$hornblende" isogrades, i.e. about 500—520 °C, which is somewhat lower than the value estimated for the gneiss — mica schist group.

Plate I

1. Textural picture of garnetiferous staurolitic muscovite-biotite schist (sample GB—29.2).

Legend: gr — garnet, st — staurolite, bi — biotite, mu — muscovite. 1 nicol, the picture width corresponds to 2 mm.

2. Syn-tectonic staurolite (st) and biotite (bi) grains in gneiss (samples So—3.3). 1 nicol, the picture width corresponds to 2 mm.

3. Post-tectonic kyanite grains (ky) in garnetiferous staurolitic kyanitic muscovite-biotite gneiss (sample Nagy—K—1.11). 1 nicol, picture width is 2 mm.

4. Staurolite and kyanite in gneiss (sample Nagy—K—11). Legend: ky — kyanite, st — staurolite, bi — biotite. 1 nicol, picture width is 2 mm.

In order to determine the physical parameters of the first progressive metamorphic event more precisely, different *geothermometric and geobarometric methods* were used. The cation numbers per unit cells necessary to the computations were calculated from the coexisting minerals analysed by means of electron microprobe (Table 6). To satisfy the requirements of the thermodynamic equilibrium, the analytical data of adjoining or close-lying mineral grains were used, the homogeneity of the chemical composition within the rock and within the individual grains was also studied. In case of minerals with chemical zoning (garnets) the composition of grain's margin was taken into account.

TABLE 6

Average chemical composition of rock forming and accessory minerals
(electron microprobe data in weight percent and numbers of cations per unit cell)
(Analysed by G. NAGY and G. DOBOSI)

1. Garnet analyses

	1.	2a	2b	3a	3b	4a	4b	5a	5b	6a	6b
SiO_2	37,84	36,57	36,89	37,63	37,80	37,17	37,32	38,40	38,43	36,77	37,41
TiO_2	0,15	0,04	0,05	n. d.	n. d.	0,06	0,03	0,19	0,12	0,21	0,23
Al_2O_3	21,39	20,84	21,17	21,49	21,40	21,37	21,35	21,21	21,64	21,10	21,11
*FeO	32,82	34,36	34,43	33,27	36,36	30,59	32,23	30,03	32,29	23,42	28,64
MnO	0,99	1,55	2,12	2,05	0,53	1,80	1,57	2,11	0,38	6,70	1,16
MgO	3,59	2,33	2,41	1,79	2,85	2,62	2,99	1,23	3,47	1.00	1,38
CaO	4,28	3,57	3,31	3,70	0,90	6,74	5,03	8,95	5,90	11,14	11,37
K_2O	0,00	0,03	0,02	n. d.	n. d.	n. d.	n. d.	n. d.	n. d.	n. d.	n. d.
Na_2O	0,01	0,03	0,00	n. d.	n. d.	0,01	0,11	n. d.	n. d.	0,03	0,03
Total	101,07	99,32	100,40	99,93	99,84	100,36	100,63	102,12	102,23	100,37	101,33

Numbers of cations on the basis of 24 O

	1.	2a	2b	3a	3b	4a	4b	5a	5b	6a	6b
Si	5,977	5,953	5.941	6,040	6,068	5,929	5,945	6,025	5,984	5,889	5.919
Al	3.983	3.998	4.018	4,065	4,049	4,017	4.009	3.922	3.971	3,983	3,938
Ti	0,017	0,005	0,005	n. d.	n. d.	0,007	0,003	0,003	0,023	0,024	0,027
$^*Fe^{2+}$	4.335	4.678	4,637	4,467	4,881	4,084	4,295	3,940	4,205	3,137	3,791
Mn	0,132	0,214	0,288	0,278	0,073	0,242	0,211	0,280	0,051	0,909	0,156
Mg	0,845	0,565	0,579	0,430	0,682	0,625	0,710	0,287	0,805	0,239	0,326
Ca	0,723	0,622	0,571	0637	0,153	0,150	0,857	1,505	0,985	1,913	1,927
K	0,001	0,011	0,008	n. d.	n. d.	n. d.	n. d.	n. d.	n. d.	n. d.	n. d.
Na	0,005	0,007	0,001	n. d.	n. d.	0,003	0,036	n. d.	n. d.	0,011	0,009
Total	16,018	16,053	16,048	15,917	15,906	16,057	16,066	15,982	16,014	16,105	16,093

1. So—3.3: gneiss, $n=3$; 2. Víz—I.5: mica schist, a) core ($n=3$) and b) edge ($n=3$) of the grains; 3. GB—29,2: mica schist, a) core ($n=3$), b) edge ($n=3$); 4. Dar—3.11: gneiss, a) core ($n=3$), b) edge (n=3); 5. Nagy—K—1.11: gneiss, a) core ($n=2$), b) edge ($n=2$); 6. Dar—Ny—2.4; amphibolite, a) core ($n=2$), b) edge ($n=2$).

* Total Fe calculated as FeO and Fe^{2+}, respectively
n = number of analyses
n. d. = not determined

In case of the gneiss and mica schist samples the complex plagioclase-biotite-garnet-muscovite thermobarometer elaborated by GHENT and STOUT [1981] was used. Results are found in *Fig. 3* and Table 7. The temperature and pressure intervals of 550—600 °C and 5.9—8.9 kbar obtained for the first progressive metamorphism correspond fairly well to the values estimated from the mineral assemblage on the basis of Winkler's reaction isograde system, as well as to the values of 510 °C and 7.3 kbar obtained for the amphibolite of Dar—Ny—2.4 by the amphibole-plagioclase thermobarometer elaborated by PLYUSNINA [1981, 1982: *Fig. 4*].

TABLE 6 (continued)

2. Plagioclase analyses

	1.	2.	3.	4.	5.	6.
SiO_2	64,14	61,57	65,40	60,05	64,57	65,63
Al_2O_3	23,14	22,32	21,31	24,09	23,27	22,72
CaO	4,57	4,77	2,13	5,88	3,89	2,85
Na_2O	9,00	8,93	10,78	8,55	9,48	9,95
K_2O	0,09	0,07	0,14	0,03	0,08	0,10
Total	100,94	97,66	99,76	98,60	101,29	101,25
	Numbers of cations on the basis of 8 O					
Si	2,806	2,791	2,886	2,708	2,812	2,903
Al^{iv}	1,193	1,192	1,108	1,280	1,194	1,111
Ca	0,215	0,231	0,101	0,284	0,182	0,126
Na	0,763	0,785	0,922	0,748	0,800	0,800
K	0,005	0,005	0,008	0,002	0,005	0,006
Total	4,982	5,004	5,025	5,022	4,993	4,946

1. So—3.3: gneiss, $n=3$; 2. Víz—I.5: mica schist, $n=3$;
3. GB—29.2: mica schist, $n=3$; 4. Dar—3.11: gneiss, $n=3$;
5. Nagy—K—1.11: gneiss, $n=3$; 6. Dar—Ny—2.4: amphibolite, $n=6$

3. Hornblende analyses
(Dar—Ny—2.4: amphibolite, $n=8$)

	Weight %		Cation numbers(23 O)
SiO_2	43,28	Si	6,480
TiO_2	0,60	Al^{iv}	1,520
Al_2O_3	12,50		
*FeO	17,52	Al^{vi}	0,686
MnO	0,27	Ti	0,068
MgO	10,68	$^*Fe^{2+}$	2.194
CaO	10,37	Mn	0,034
Na_2O	2,03	Mg	2,448
K_2O	0,29		
Total	97,54		
		Ca	1,664
		Na	0,589
		K	0,054
		Total	15,737

* Total Fe calculated as FeO and Fe^{2+}

4. Biotite analyses

	1.	2.	3.	4.	5.
SiO_2	36,10	34,81	36,16	39,07	35,97
TiO_2	1,78	1,79	1,83	1,71	1,88
Al_2O_3	20,99	17,93	19,63	18,87	19,84
*FeO	16,56	22,07	20,35	15,77	16,78
MnO	0,11	0,10	0,17	n. d.	0,05
MgO	10,94	8,37	8,92	11,56	11,71
CaO	0,00	0,04	n. d.	n. d.	0,06
Na_2O	0,23	0,06	0,36	0,19	0,19
K_2O	9,17	8,93	8,91	8,93	8.19
Total	95,88	94,10	96,33	96,10	94,67

Numbers of cations in the basis of 22 O

Si	5,355	5,437	5,439	5,700	5,389
Al^{iv}	2,645	2,563	2,561	2,300	2,611
Al^{vi}	1,024	0,737	0,918	0,967	0,893
Ti	0,198	0,211	0,207	0,188	0,212
$*Fe^{2+}$	2,055	2,884	2,559	1,935	2.105
Mn	0,015	0,013	0,022	n. d.	0,007
Mg	2,421	1,949	2,001	2,526	2,615
Ca	0,000	0,006	n. d.	n. d.	0,009
Na	0,067	0,019	0,104	0,054	0,055
K	1,736	1,780	1,709	1,669	1,565
Total	15,516	15,599	15,520	15,339	15,461

1. So—3.3: gneiss, $n=2$; 2. Víz—I.5: mica schist, $n=7$;
3. GB—29.2: mica schist, $n=5$; 4. Dar—3.11: gneiss, $n=3$;
5. Nagy—K—1.11: gneiss, $n=4$.

* Total Fe calculated as FeO and Fe^{2+}

n. d. = not determined.

5. Muscovite analyses

	1.	2.	3.	4a	4b	5.
SiO	44,50	45,85	46,93	45,76	46,20	46,34
TiO_2	0,40	0,51	n. d.	0,61	0,46	0,51
Al_2O_3	34,81	33,57	34,94	34,21	33,86	34,74
*FeO	0,65	1,15	0,83	1,07	1,85	1,20
MnO	0,02	0,01	n. d.	0,04	0,01	0,01
MgO	0,60	0,86	0,66	1,06	1,46	0,91
CaO	0,00	0,02	0,05	0,01	0,00	0,01
Na_2O	1,16	1,09	1,56	1,06	0,39	1,23
K_2O	9,59	9,64	9,45	9,69	9,95	9,47
	91,73	92,70	94,42	93,51	94,18	94,42

	Numbers of cations on the basis of 22 O					
Si	6,125	6,224	6,260	6,190	6,220	6,198
Al^{iv}	1.875	1,776	1,740	1,811	1,780	1,802
Al^{vi}	3,772	3,656	3,756	3,644	3,593	3,674
Ti	0,041	0,053	n. d.	0,063	0,046	0,052
$*Fe^{2+}$	0,074	0,132	0,093	0,122	0,208	0,135
Mn	0,003	0,001	n. d.	0,005	0,002	0,001
Mg	0,124	0,176	1,131	0,214	0,293	0,182
Ca	0,000	0,003	0,008	0,001	0,001	0,002
Na	0,308	0,289	0,403	0,278	0,103	0,320
K	1,684	1,687	1,608	1,672	1,709	1,617
Total	14,006	13,997	13,999	14,000	13,955	13,992

1. So—3: gneiss, $n=2$; 2. Viz—I.5: mica schist, $n=4$;
3. GB—29.2: mica schist, $n=3$; 4. Rul—3.2: phyllonitic blastomylonite, a) large porphyroclasts, $n=2$, b) small sericite flakes, $n=3$; 5. Nagy—K—1.11: gneiss, $n=8$.

* Total Fe calculated as FeO and Fe^{2+}
n. d. = not determined

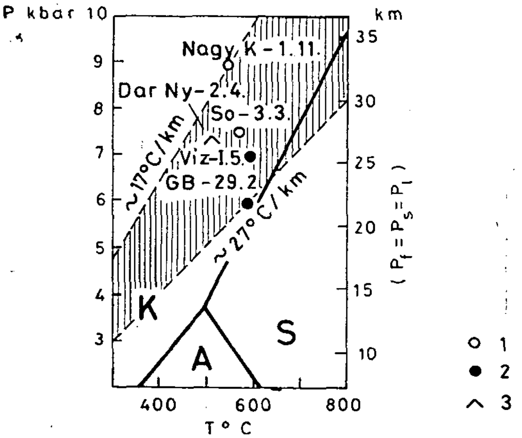


Fig. 3. P—T conditions and geothermal gradient-range of the first progressive metamorphism of the crystalline basement based on the geothermobarometer of GHENT and STOUT [1981] showing also the values obtained by the method of PLYUSNINA [1981 1982] for amphibolites. Legend: 1 — gneiss, 2 — mica schist, 3 — amphibolite, A — andalusite, K — kyanite S — sillimanite.

This latter thermobarometer is based on the P—T dependent changes of the anorthite content of the plagioclase and of the Al-content of hornblende in zoisite-epidote containing assemblages.

Presuming the pressure relations $P_f = P_s \approx P_l$ (which is generally accepted for the amphibolite facies conditions, [see WINKLER, 1976, 1979] and using an average rock density value, the pressure — depth — temperature relationship is demonstrated in *Fig. 3* (P_f — fluid pressure, P_s — pressure acting on the solid phases and P_l — lithostatic or load pressure). Accordingly, the first progressive metamorphism was going

TABLE 7

Temperature and fluid pressure conditions of the first progressive metamorphic event

Sample	Rock type	Thermobarometers			
		plagioclase-biotite garnet-muscovite (Ghent and Stout, 1981)		amphibole-plagioclase (Plyusnina, 1981, 1982)	
		T [°C]	P_f [Kbar]	T [°C]	P_f [Kbar]
So—3.3	sillimanite-staurolite-garnet-biotite-muscovite-gneiss	564	7,5	—	—
Víz—I.5	garnet-biotite-muscovite schist	598	6,9	—	—
GB—29.2	staurolite-garnet-muscovite-biotite schist	582	5,9	—	—
Dar. Ny—2.4	garnet-epidote-biotite amphibolite	—	—	510	7,3
Nagy—K-1.11	kyanite-staurolite-muscovite-biotite gneiss	551	8,9	—	—

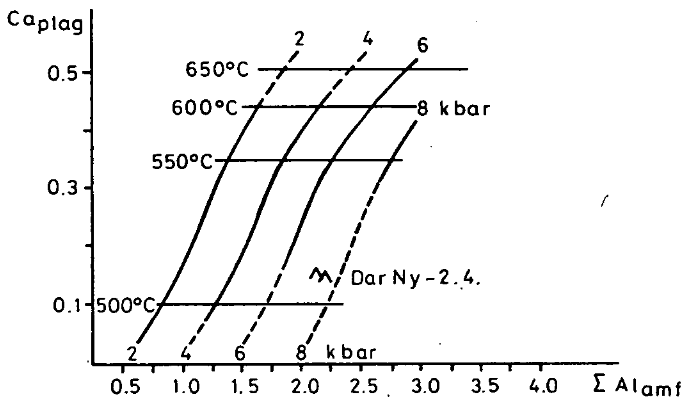


Fig. 4. P—T conditions of amphibolite metamorphism according to the geothermobarometer of PLYUSNINA [1981, 1982] for amphibole-plagioclase.

on in a system with a geothermal gradient interval of 17—27 °C/km. This result supports the former statement of LELKES-FELVÁRI, SASSI *et al.* [1981] giving the first quantitative data on the geothermic regime of the metamorphism. This gradient interval is characteristic of the medium pressure range (Barrovian) facies series. The garnet, staurolite, kyanite, and according to the thermobarometric analysis of the sample So—3.3, the sillimanite were generated by this metamorphism.

Traces of anatectic or potash-metasomatic migmatite formation could not be evidenced in the samples except a few uncertain cases. Nevertheless, the plagioclase (oligoclase) porphyroblasts are common and characteristic in the rocks. The formation of plagioclase porphyroblasts (more precisely: poikiloblasts containing the often resorbed inclusions of quartz, biotite, muscovite and garnet, see Plate II.2) together with the rarely observable mirmekite formation can be evaluated as a phase directly preceeding the anatexis in the course of the recrystallization.

The presence of andalusite in one sample indicates that in addition to the Barrovian metamorphic event an other (low-pressure) amphibolite facies meta-morphism of >34 °C/km gradient also acted in certain, restricted parts of the Somogy—Drava Basin. The relationship between andalusite and kyanite is unknown. As in the studied sample of Kut—2.16 the andalusite is younger than the mostly syntectonic garnet+staurolite assemblage (the andalusite contains resorbed staurolite inclusions, see Plate II.1), it can be assumed that the medium pressure kyanitic type recrystallization was overprinted by a low-pressure andalusite-type metamorphism in a subsequent tectonophase or -cycle. (Because of the weathered state of the anda-lusite-bearing sample the thermobarometric method could not be used.)

Presumably this second (andalusitic) metamorphic event can be related to the Hercynian granitization of the Mecsek Mountains and of the Danube—Tisza Interfluve from spatial, temporal and genetic points of view as well.

The two metamorphic events outlined above were followed by a younger, low-temperature retrograde metamorphism. Its temperature did not reach the biotite isograde (~ <450 °C, see WINKLER, 1976, 1979) and might be 200—400 °C corre-sponding to the anchizone and the low temperature part of the greenschist facies (chlorite-zone, epizone). Its mineral assemblages consisting of quartz, sericite, chlorite, calcite, siderite and dolomite can not be or can only be hardly distinguished from the products of the subsequent weathering processes. The illite crystallinity parameters of the not weathered or slightly weathered, biotite-free retrograde meta-morphic samples indicate mostly epizonal (greenschist facies) conditions. The fluid pressure of the retrograde metamorphism can be estimated only on the basis of the b_0 average of the white micas ($\bar{x}$=8.996 Å, n=3) which presumes low-pressure con-ditions. The carbonate minerals being always present in these assemblages indicate that the CO_2 in addition to the H_2O might have played an important role in the fluid system of the retrograde recrystallization.

Locally, the retrograde metamorphism was connected (most probably somewhat preceeded) by cataclastic metamorphism (mylonite formation): samples Rül—2.2, Rul—3.2, GB—29.4, Se—2.18, Kkut—2.4—9 (Plate II.4). The mineral assemblages of the "static type" retrograde metamorphism characterized above and the assem-blages of the weak matrix-recrystallization (blastomylonite formation) subsequent to the dynamic mylonitization are the same. There is a considerable difference in the phengite content of muscovite generated by the first progressive metamorphism and of sericite generation produced by blastomylonite formation (see Table 6 point 5).

When summing up the conclusions the *evolution of the metamorphic basement* can be divided in the phases below, in relative time scale:

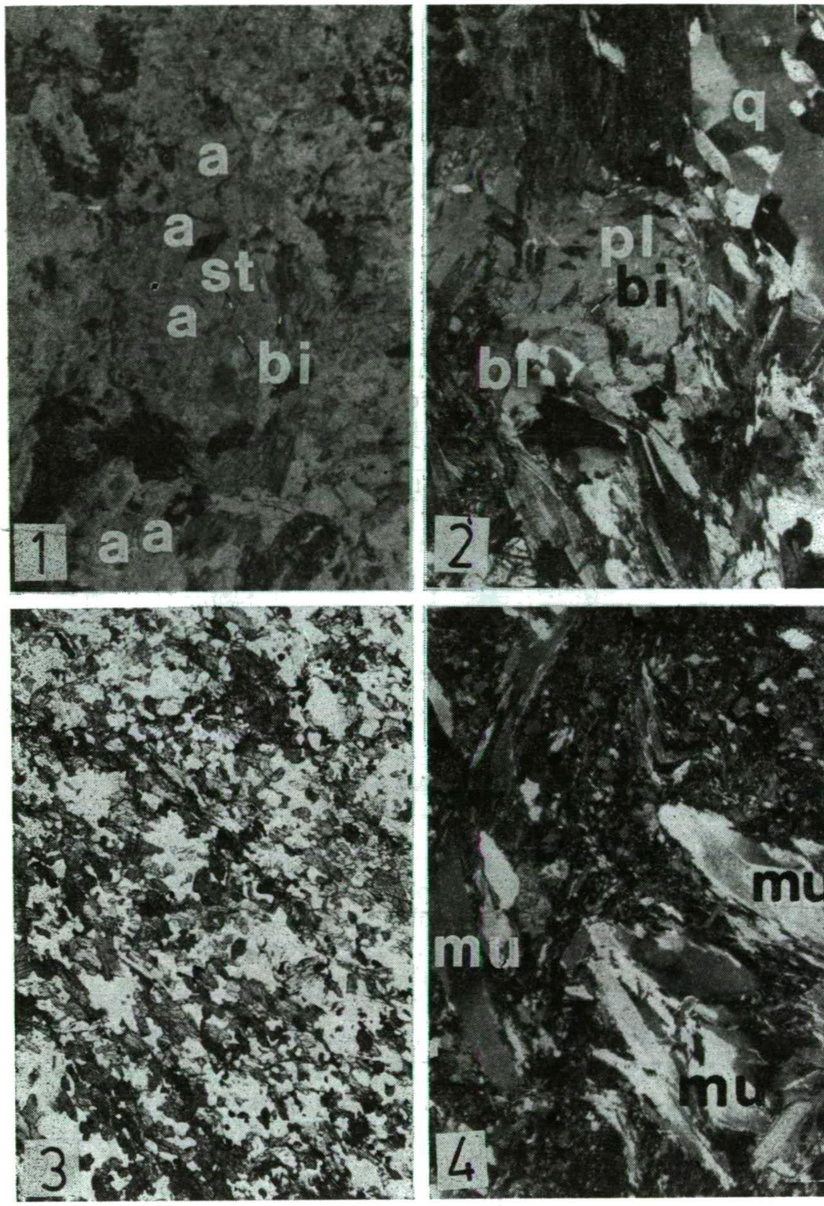
A) the oldest event was a Barrovian medium pressure, medium-grade (alman-dine-amphibolite facies) regional metamorphism with a geothermal gradient range of 17 to 27 °C/km. This was followed by

B) and andalusite-type (low-pressure, >34 °C/km gradient) amphibolite facies overprint, and

C) a predominantly low-pressure, low-temperature, anchi-, epizonal regional metamorphism (<450 °C), locally with cataclastic (mylonitic, blastomylonitic) character.

By means of geothermometric and geobarometric methods only the physical parameters of the event *A)* could be determined: 510—600 °C and 5.9—8.9 kbar.

PLATE II



Having no isotope geochronological data and counting with the chronological uncertainties of the metamorphic and granitoid rocks of the Mecsek Mountains adjoining to the investigated area, the chronological classification of the metamorphic events can be made on the bases of farther and presumed analogies. Consequently, several hypothetic models can be established. The possibilities exemplified by the Eastern Alps (FRANK, PURTSCHELLER *et al.*, in NIGGLI 1978] are as follows:

1) *A)* Caledonian, *B)* Hercynian and *C)* Hercynian and/or Alpine:

2) *A)* older Hercynian, *B)* younger Hercynian and *C)* younger Hercynian and/or Alpine.

Based on the models of the Carpatho-Pannonian region [SZÁDECZKY–KARDOSS, ÁRKAI *et al.*, 1976]:

3) *A)* Dalslandian = Early Baikalian, *B)* Hercynian and *C)* Hercynian and/or Alpine.

Out of these models the varieties 1 and 2 seem to be the most probable. To prove them, however, further detailed and systematic isotope geochronological investigations are required.

WEATHERING

By means of the usual microscopic investigations the retrograde metamorphic mineral assemblages consisting of sericite, chlorite, quartz and carbonate minerals can not be distinguished from the similar mineral parageneses formed by surficial or near-surface chemical weathering of the crystalline basement rocks by the migration of low-temperature solutions. Thus, concerning the effects of weathering conclusions could be drawn only from the qualitative and quantitative relations of clay minerals forming under low-temperature conditions, and from the illite crystallinity indices of the biotite-free samples.

Based on the clay mineral assemblages (Table 2) it can be stated that in the samples investigated the degradation process being progressively parallel with increasing weathering was incomplete. Out of the members of the illite→kaolinite→→mixed layer clay mineral→smectite degradation (weathering) series determined by RIEDMÜLLER [1978] for phyllosilicate bearing metamorphic rocks only kaolinite and illite could be detected in the crystalline basement of the Somogy—Drava Basin. Out of the samples investigated, ten samples contained kaolinite between trace and

Plate II

1. Andalusite (a) with inclusions of staurolite (st) and biotite (bi) in gneiss sample Kut—2,16. 1 nicol, picture width is 1 mm.

2. Plagioclase poikiloblast (pl) containing resorbed biotite (bi) inclusions in mica schist (sample GB—29.2). Legend: q — quartz. Crossed nicols, picture width is 2 mm.

3. Textural picture of amphibolite (sample Dar—Ny—2.4). 1 nicol, picture width is 2 mm.

4. Textural picture of blastomylonite of gneissic origin with muscovite (mu) porphyroclasts. — crossed nicols, picture width is 2 mm.

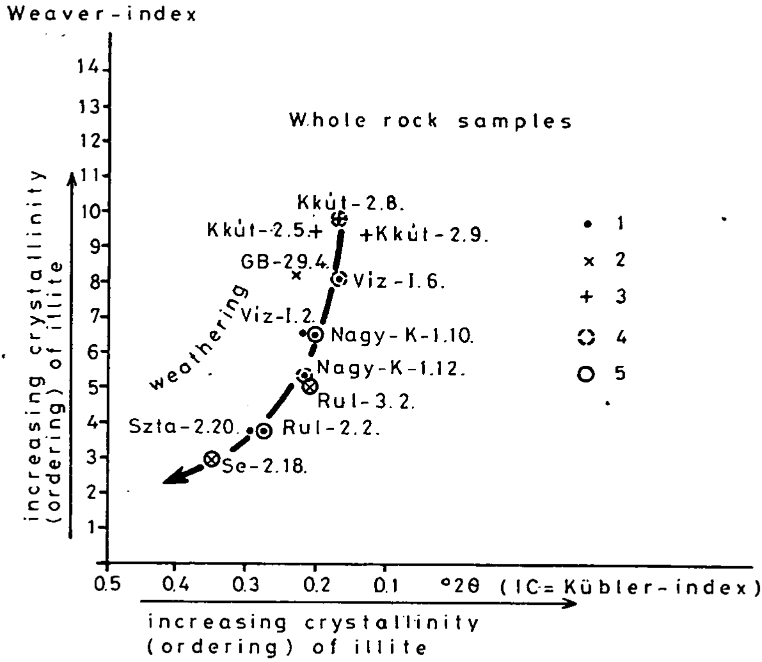


Fig. 5. Effect of weathering on the illite crystallinity indices of the biotite-free metamorphic rock Legend: 1 — gneiss and mica schist, 2 — mylonite, 3 — blastomylonite, 4 — trace — 5% kaolinite 5 — >5% kaolinite content.

5%, and five samples between 5 and 15%, thus regarding the quality, the chemica weathering was partial and initial, and concerning the quantitative relations, only o minor significance.

In the samples not containing biotite the illitization producing low-temperature disordered lattice can be detected only by the illite crystallinity indices, first of all by the Weaver-index. The results of this unusual, novel application of illite crystallinity methods used so far only to distinguish the progressive stages of the diagenesis and incipient metamorphism, are shown in *Figs. 5* and *6.*

It can be seen in *Fig. 5* that in the initial stage of weathering (when the samples contain small quantities of poorly crystallized, disordered illite) the Weaver-index considerably decreases while the Kübler-index remains nearly the same as in the not weathered samples. This phenomenon can be explained by a mixture of the muscovite having good crystallinity and of illite with not perfect, disordered structure. The increasing illitization is usually accompanied by the appearance of kaolinite and by its increasing amount. A tendency similar to that observed in the bulk rock samples can be observed in the fractions of less than 2 microns too *(Fig. 6).*

It can be also assumed that at least a part of the chlorite was generated also by weathering. Its proof, however, needs further investigations.

The carbonate minerals generated by retrograde metamorphism and formed also in the subsequent surficial or near-surface weathering can not be distinguished either. The carbonate formation took place in several phases from the diffuse imbibition up to the fissure filling with sharp boundaries producing the formation of calcite, dolomite and siderite.

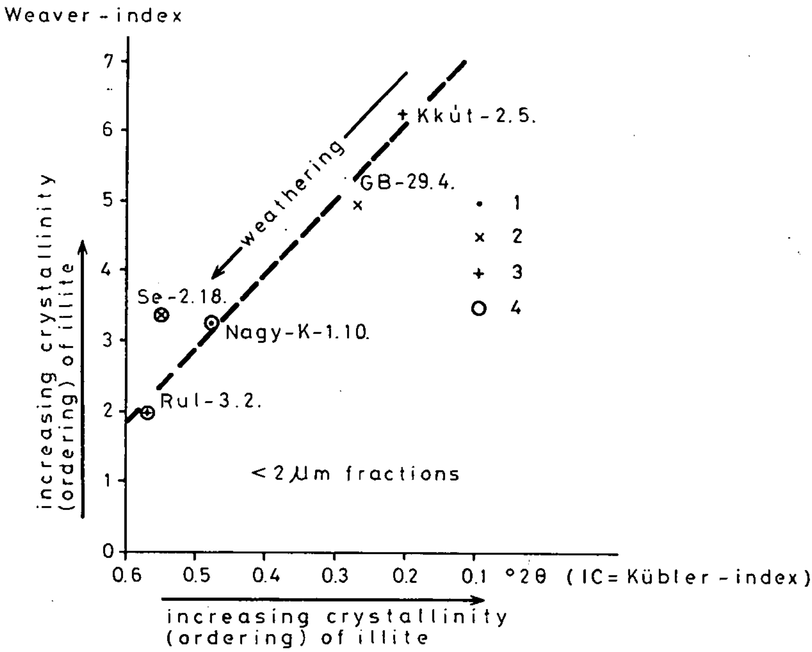


Fig. 6. Effect of weathering on the changes of illite crystallinity of the fractions of less than 2 microns in relation with the kaolinite content of the whole rock samples.
Legend: 1 — gneiss, mica schist, 2 — mylonit,e 3 — blastomylonite, 4 — >5% kaloinite content.

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P. ÁRKAI
Laboratory for Geochemical Research
Hungarian Academy of Sciences
H-1112 Budapest, Budaörsi út 45
Hungary